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TESTING THE GEOLOGICAL INFLUENCES ON THE EVOLUTION OF RIVER PROFILES: A CASE FROM THE NORTHERN APENNINES (ITALY)

ABSTRACT: SPAGNOLO M. & PAZZAGLIA F.J., *Testing the geological influences on the evolution of river profiles: a case from the northern apennines (italy)*. (IT ISSN 1724-4757, 2005).

Longitudinal river profiles of the Emilia and Romagna Apennines rivers were modeled using a slope-area and stream power law approach in order to verify the influence of variable rock types and hypothesized higher rates of rock uplift in the Romagna region. In particular, long river profiles were studied in terms of their concavity and steepness. A sensitivity analysis using different methods of sampling a DEM and subsequent profile smoothing established the details of our methodology. We find concavity and steepness both to be affected by the sampling method which subsequently guides our selection of 14 long trunk rivers and 15 small headwater channels distributed evenly between the Emilia and Romagna portions of the range. Modeled concavity and steepness values are largely consistent with those reported by similar studies and lie close to the stream power law predicted values for steady-state channels. Overall, profile concavity is higher in the Emilia portion, especially for the small channels, with respect to the Romagna region. In both the main trunks and small channels analyses, steepness is also higher in the Emilia region. These results have three possible interpretations. The rocks in the Emilia Apennines offer greater overall resistance to erosion than those in Romagna, although the traditional notion is that the siliciclastic turbidites of Romagna should be more durable than the Ligurian units of Emilia. Alternatively, rock uplift in Emilia is currently greater than in Romagna, but this observation stands in contrast to the fact that the Ligurian rocks are still intact as a structural lid in Emilia. Lastly, the results may be interpreted as sediment flux and comparative rate of down stream fining in the channels with the Emilia rivers characterized by coarser grain sizes, especially in the headwaters region. The last possibility makes a prediction about the variable hydrology of watersheds underlain by variable rock type that can be directly tested as a follow up to this study.

KEY WORDS: Fluvial geomorphology, Landscape evolution, Northern Apennines, Longitudinal river profiles, Slope vs. area analysis.

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INTRODUCTION

Tectonic geomorphology is predicated on successfully isolating the influence of rock deformation from rock-type, climate, and non-tectonic base level changes to correctly interpret landscapes in the context of constructive tectonic processes. The proliferation of high resolution digital elevation models has fueled the rapid extraction of landscape metrics that are traditionally viewed as key indicators of active tectonics (Mayer, 2000). Of these metrics, river longitudinal profiles have long been viewed as sensitive indicators of the rate of rock uplift as the rate of river incision is typically assumed to be driven by the long-term rate of local or regional base level fall (Ohmori, 1991; Knuepfer, 1994; Pazzaglia & *alii*, 1998). Despite efforts to uncover the differences between long profiles with the hopes of extracting a tectonic signal, most river long profiles typically assume a similar concave-up shape, defined by a characteristic concavity, regardless of the tectonic setting or rate of rock uplift (Snyder & *alii*, 2000; Kirby & Whipple, 2001). This concave-up shape, found for both alluvial as well as bedrock channels, is a direct result of a downstream increases in discharge and decreases in grain size and speaks perhaps to the overall hydrologic and sediment transport similarities among diverse watersheds, rather than their tectonic differences. Correct interpretation of tectonic processes from subtle changes in long profile shapes therefore hinges on identifying those characteristics of the long profile that are sensitive to tectonics and directly comparable across watersheds of variable size. Recent studies have made progress in this direction suggesting that the overall gradient of a stream long profile, a quantity called steepness, may be sensitive to uplift rate (Hurtrez & *alii*, 1999; Snyder & *alii*, 2000, 2003; Kirby & Whipple, 2001). Other, non-tectonics effects on long profile curvature or concavity, including rock-type (Hack, 1957) and climate are typically acknowledged, but rarely resolved in most contemporary studies (c.f. Snyder & *alii*, 2000, 2003; Roe & *alii*, 2002; Duvall & *alii*, 2004). Our goal in this paper is to specifically consider the influence of rock-type on long profile form and attempt to isolate its effect from that of rock uplift in a tectonically active setting where rivers are incising into bedrock at rates between 0.2 and 2 mm/yr. Our approach is similar to that outlined in Duvall & *alii* (2004), but differs in the sense that we use both large and small channels distributed across a well-established rock-type boundary. The study is located in the Apennines of northern Italy and our results are directly applicable to broadening the understanding of several competing models that have recently been proposed for the emergence of topography and long term landscape evolution in this orogen.

River incision into bedrock

The longitudinal profile (long profile) of a river is a plot of the channel elevation with respect to distance. The valley long profile, which is a plot of the medial valley elevation with respect to distance projected to a vertical plane bisecting the long axis of the valley is a better average reflection of the long-term long profile responsible for carving the valley. The rate at which a long profile flattens downstream is termed the profile concavity. Profile concavity tends to vary less in diverse tectonic settings in comparison to the overall relief of the profile from its headwaters to the mouth. Once a stream reaches a concavity dictated by discharge, grain size, substrate, and channel width, the stream has reached grade (Mackin, 1948) and that concavity tends to be conserved as a characteristic form (Bull, 1979) over graded time scales (Schumm & Lichty, 1965), even under variable rates of tectonic uplift (Kirby & Whipple, 2001). In contrast, profile steepness does tend to be sensitive to rates of rock uplift and varies as a power law to the rate of rock uplift (reviewed in Duvall & *alii*, 2004).

Long profiles of alluvial, mixed alluvial-bedrock, and bedrock streams all typically exhibit the concave-up shape, a result of detachment limited erosion processes on the bedrock channel and the downstream trends of grain size fining, increase in discharge, and increase in channel width in the alluvial channels (Leopold & *alii*, 1964; Sinha & Parker, 1996; Ellis & *alii,* 1997). Modeling the rate of incision of a bedrock channel has arisen through two different, but related approaches (Howard & *alii*, 1994). One approach considers the incision rate proportional to the stream power, which can be defined as the rate of energy expenditure by the flow (Seidl & Dietrich, 1992). The other approach considers incision rate proportional to the average shear stress on the bed (Howard, 1971; Howard & Kerby, 1983, from their study on badland channels). Both approaches lead to a similar equation («the stream power law», Sklar & Dietrich, 1998) for the incision rate:

$$E = K\, A^{m}\, S^{n} \qquad (1),$$

where E is the bedrock erosion rate, K is a dimensional constant that varies with rock type, climate, channel width, channel hydraulics and sediment load, A is the drainage area, usually considered a good proxy for discharge, S is the channel gradient, proportional to discharge as well. Equation 1 is valid only assuming detachment-limited processes, steady, uniform flow, a linear (or nearly linear) relationship between discharge and drainage area, a negligible threshold for activating erosion/transport and a channel width that grows as a function of the square root of the discharge. Rearranging equation (1) and solving for S, it is possible to obtain the following equation:

$$S = (E/K)^{(1/n)}\, A^{-(m/n)} \qquad (2).$$

When the m/n ratio is equal to 1, then the erosion rate is proportional to the total stream power; when m/n is equal to 0.5, the erosion rate could be still proportional to stream power per unit bed area, if channel width varies with the square root of drainage area, or to the basal shear stress at the streambed (Whipple & Tucker, 1999). Equation (2), in the case of a steady-state topography in which E is always equal to the local rock uplift rate U, can be rewritten as:

$$S = (U/K)^{(1/n)}\, A^{-(m/n)} \qquad (3).$$

Equation (3) is the simple form of a straight line in log space where the $(U/K)^{(1/n)}$ component is the y-intercept and (m/n) is the slope of the line.

Many studies (Hack, 1957; Flint, 1974; Moglen & Bras, 1995; Slingerland & *alii*, 1998; Hurtrez & *alii*, 1999; Snyder & *alii*, 2000; Kirby & Whipple, 2001) have already shown that the channel slope along a river tends to decrease inversely with the increase of the drainage area, defined by a hyperbolic equation («Flint's law», Hurtrez & *alii*, 1999) of the form:

$$S = k_s\, A^{-\theta} \qquad (4),$$

with θ being the concavity index and k_s the steepness index.

There is a symmetry between the observationally-derived equation (4) and the theoretically-derived equation (3) for the case of detachment-limited, steady state channels. If equations (3) and (4) are equivalent, θ becomes a proxy for m/n and k_s for $(U/K)^{(1/n)}$. Many authors have investigated, although not always using precisely the same methods, the variation in q and k_s where the uplift rate (channel incision rate) is independently known. For example, Hurtrez & *alii* (1999) could not find a clear relationship between the steepness index (k_s) and rock uplift in the Siwalik Hill region in Nepal, but in the same area Kirby & Whipple (2001) argue that steepness varies in dependence of rock uplift and that concavity varies for channels that are oriented parallel to a strong rock uplift gradient, whereas concavity is constant for channels that experience the same rate of rock uplift throughout the entire basin. Snyder & *alii* (2000) found a general correlation between

k_s and rock uplift in the Mendocino Triple Junction area, a prediction that was first proposed by a numeric model not specifically applied in the California Coast Ranges (Willgoose, 1994). Also θ, whose value usually varies in the range 0.4-0.7 (Tucker & Whipple, 2002), can be influenced by some factors. Specifically, it may be related to orographic precipitation distribution within a basin (Roe & *alii*, 2002), and in particular with mean annual rainfall intensity and mean peak annual discharge (Zaprowski & *alii*, 2005). Finally, given that k_s includes the K of equation (1), bedrock erodibility (rock type) should also be considered a possible influencing factor in the variability of channel steepness in actively uplifting settings. Stock & Montgomery (1999), focusing on several sample areas worldwide, found that K can vary over several orders of magnitude depending on rock type. Duvall & *alii* (2004) explore some effects of rock type on long profiles for relatively short (≤ 12 km) channels and conclude that concavity and steepness are enhanced for watershed characterized by resistant rocks and detachment limited channel processes in the headwaters and soft rocks and transport limited channel processes in the lower portion of the basin.

In this context, we perform a slope-area analysis on the long profiles of a wide range of rivers on the northeastern flank of the northern Apennines, focusing on two main goals: (1) to perform a sensitivity analysis using different methods of sampling a DEM and subsequent profile smoothing, and (2) to investigate the spatial correlation of variations in profile steepness and/or concavity primarily in terms of rock type, and secondarily in terms of climate or variations in the rock uplift rate.

STUDY AREA

We analyze channel long profiles (fig. 1) along the northeast flank of the Emilia (westward) and Romagna (eastward) Apennines, Italy. The highest elevations (1500-2000 m) lie generally along the main divide separating rivers flowing transverse to the orogen towards the Adriatic Sea from rivers flowing towards the Tyrrhenian Sea; however, the highest point is Monte Cimone (2165 m) which is situated northeast of the drainage divide. The study area covers some 11,000 km^2, has a mean elevation of 500 m, and a mean hillslope gradient of 12.3°. Lower-lying areas near the Po Plain are dominated by anthropogenic activities that are concentrated in the river valleys. The mountainous areas, after millennia of intense grazing and foresting, are now mostly protected and covered by dense tree vegetation. Nevertheless, especially where the rock type is weak and along the main river valleys such as in the northwestern portion of the study area, the area is known for its numerous very large landslides (i.e. the Corniglio landslide in the Parma Valley).

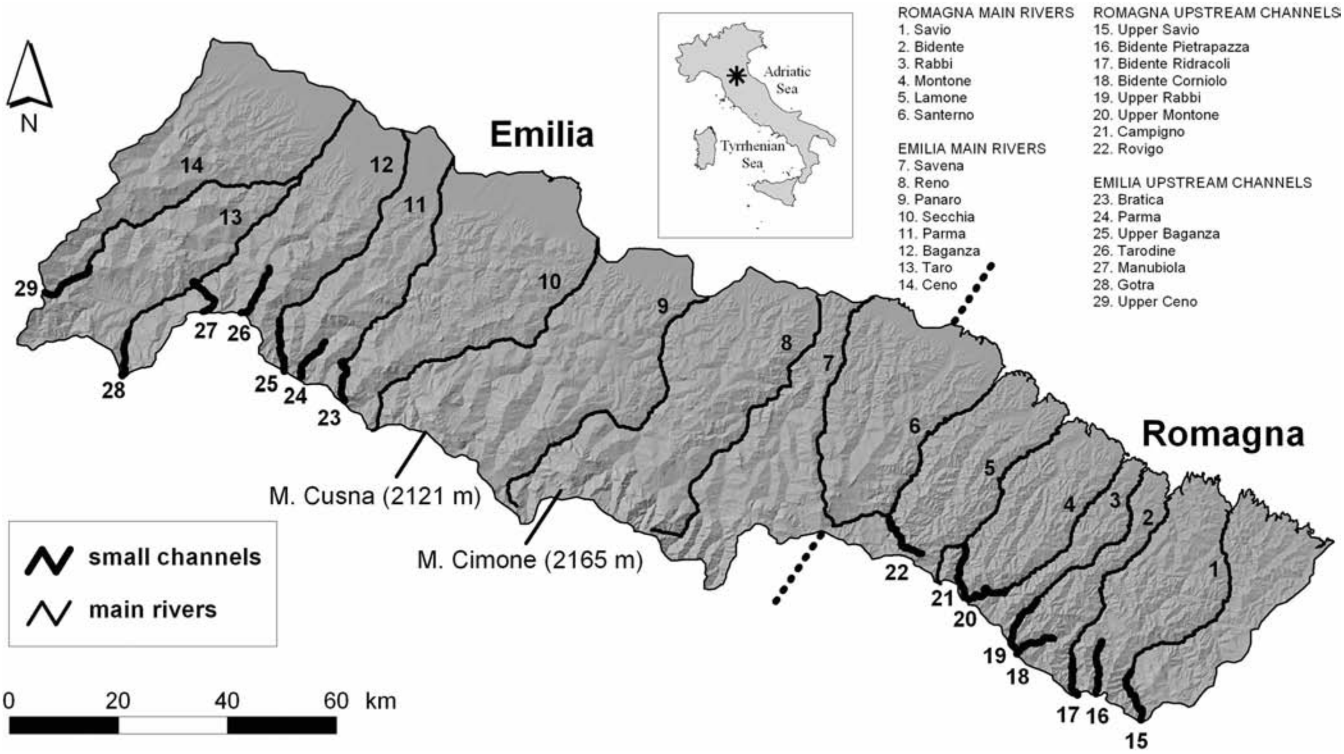

FIG. 1 - The Northern Apennine study area, with the 14 main rivers analyzed and the 15 upstream channels (numbers refer to table 2); the dotted line approximately corresponds to the Sillaro line which divide the two studied sectors.

The Northern Apennines represent concurrent Cenozoic crustal shortening and extension during subduction of the Adriatic plate. The Emilia-Romagna portion of the northern Apennines represents that part of the orogen that is still shortening, with a deformation front buried beneath Quaternary sediments of the Po foreland. Rocks in the uplifted and exposed portion of the orogenic wedge are Mesozoic carbonates and Cenozoic, mostly turbiditic sandstone, siltstone, mudstone, and marl deposited in shelf-slope-trench basins in front of and atop a thrust imbricated Mesozoic platform carbonate basement. These siliciclastic rocks have now been frontally accreted and incorporated into the wedge. In the Emilia part of the range (fig. 2), the Cenozoic turbidites remain buried beneath a largely intact structural lid called the Ligurian nappe which represents a large thrust panel composed of Mesozoic ophiolite basement, Mesozoic marine siliciclastics and carbonates, and epi-Ligurian (wedge-top), shelf-slope siliciclastic basins. In contrast, in the Romagna part of the range the Ligurian structural lid has been removed, presumably by subareial erosion since the Messinian, revealing the underlying Oligo-Miocene foredeep deposits of the Marnoso-Arenacea Formation (Cerrina Feroni & *alii*, 2001). The estimated tectonic overload of Ligurian nappe and epi-Ligurian successions for the Romagna part is about 4500 m which is consistent with the 4000 m currently cropping out in the Emilia (Cerrina Feroni & *alii*, 2001). There is an abrupt contrast in relative rock strength, degree of stratification, and structural deformation between the Emilia and Romagna Apennines separated more or less along the valley of the Sillaro River (the Sillaro Line on fig. 2). In general, the Romagna siliciclastics are thinly bedded, stratigraphically and structurally coherent, and locally contain laterally extensive resistant units. Landslides are not common here and steep cliffs persist among the resistant strata. The various Ligurian rocks, in contrast, are highly tectonized, lack lateral coherence, and dominated by mostly incompetent mudstone. Resistant rock types, such as limestone or chert, occur as isolated bodies, often olistrostomic, and as a result the entire region is prone to large-scale mass movements. Rock resistance is generally lower in the Emilia Apennines in comparison to the Romagna Apennines except for some of the high elevation portion near the drainage divide. Here, siliciclastic turbidites, the Macigno and Cervarola formations, similar in grain size and bedding characteristics to the younger Marnosa Arenacea, outcrop as resistant cliff-formers. At the large scale on satellite images (Landsat TM), the Romagna Apennines are classified as much less rough in terms of morphological texture than the Emilia Apennines (Bartolini & Carton, 1992).

The climate for both the Emilia and Romagna Apennines is similar, characterized by a continental, seasonal distribution of precipitation. Mean annual temperature falls and mean annual precipitation rises along a northeast to southwest gradient parallel to the rise in mean elevation. From a study that took into account data collected from

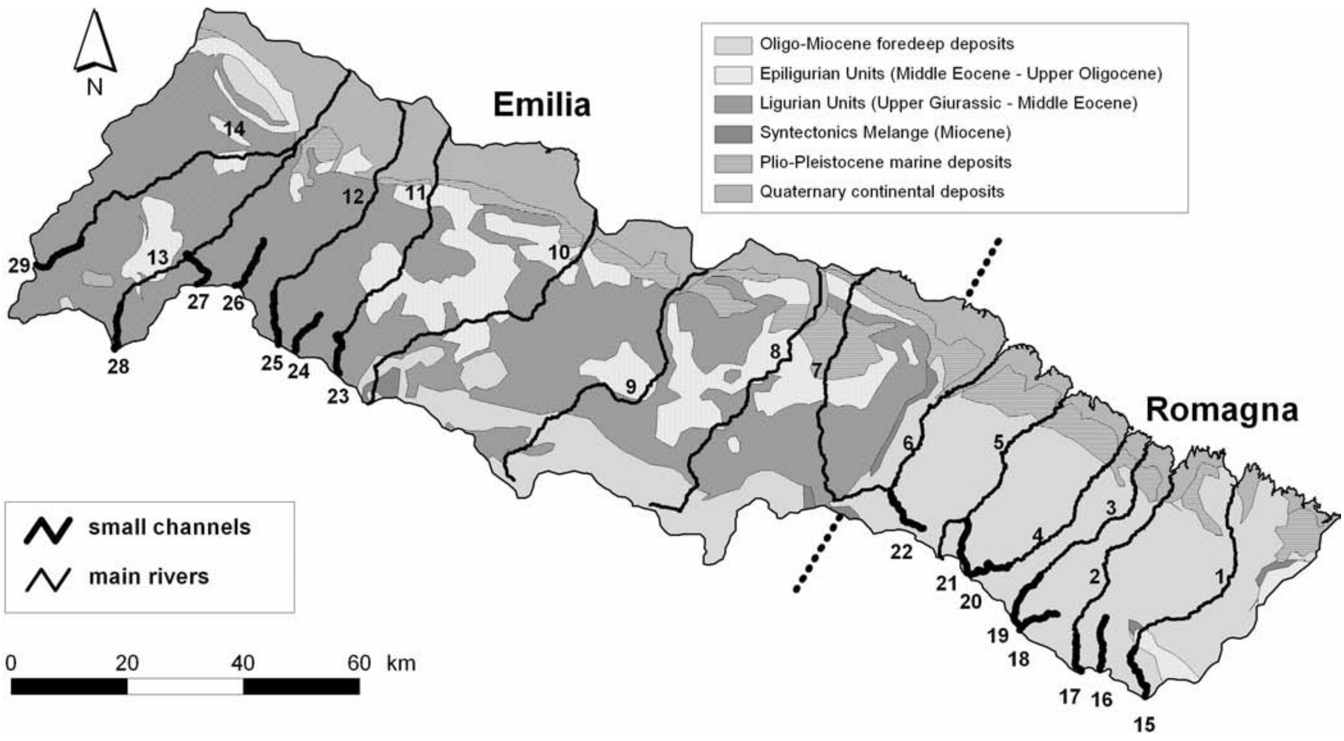

FIG. 2 - A lithologic schematic map of the area with the 14 main rivers analyzed and the 15 upstream channels (numbers refer to table 2); the dotted line approximately corresponds to the Sillaro line which divide the two studied sectors.

several meteorological stations for more than 40 years (Regione Emilia-Romagna, 1995), it is possible to infer that, all over the studied area, mean annual temperature varies from 12 to 8 °C and mean annual precipitation from 800 to 1500 mm, mostly depending on the altitude. There are sparse evidences of glaciation during the Pleistocene with rare glacial deposits, like the ones found at an elevation less than 800 m asl in the Val Parma, where the reconstructed snowline was at least 1260 m high (Federici & Tellini, 1983); however, there are no Holocene glaciers or permanent snow fields.

Major drainages flow transverse to the orogens in long, generally linear valleys. Headwaters have a dendritic pattern whereas trellis patterns are more typical in the lower reaches of the watersheds proximal to the Po Plain. Most of the rivers in the Romagna Apennines, including and in between the Rubicone River and the Reno River, flow directly into the Adriatic Sea. These rivers have bedrock channels all the way from their headwaters to the mountain front where they make the transition to alluvial channels on the Po Plain. In the Emilia Apennines from the Panaro River to the Baganza River the rivers are all tributary to the Po River, the main, axial drainage of the Po foreland basin. The lower quarter of these Emilian channels are alluvial, but the upper three quarters have channels on bedrock. We restrict our analysis to only that portion of the rivers upstream of the Apennine mountain front which are dominantly detachment-limited bedrock channels.

METHODS

We apply the slope-area analysis to three nested spatial-temporal levels (figure 1): (1) to one single river (the Secchia River), central in the study area and with a distinctive, relatively smooth concave-up profile, good for calibrating our methods, (2) to the main 14 rivers in the study area, mainly bedrock channels of Strahler order equal to or higher than six, and a length measured from the headwaters of approximately 60 km and (3) to 15 selected bedrock streams near the divide with uniform lithology distributed equally between Ligurian and Marnosa Arenacea rock types of Strahler order equal to or lower than 4 and lengths measured from the headwaters of approximately 10 km.

Long profiles were extracted from a 10 m horizontal resolution digital elevation model derived from topographical maps at a scale of 1:5,000. The DEM was first processed to remove data errors including sinks. Hydrologic GIS analysis were applied using ArcView 3.2 in order to obtain flow direction and flow accumulation maps of the study area as well as the drainage network. A minimum flow accumulation (or drainage area) threshold of 0.02 km^2 was applied to define the headwater extent of first-order channels. The minimum drainage area threshold considered here is very low compared to what would be commonly used to extract the drainage network for fluvial pattern analysis or for generic quantitative fluvial geomorphology studies. Nevertheless, the 0.02 km^2 drainage area threshold was applied with the specific intent of testing whether the gradient of small channels, or perhaps even colluvial hollows in the drainage network would affect the river profile analysis.

We extract the trunk channel from an ordered drainage network of the entire basin, including tributaries. The trunk channel is defined as the longest possible path moving upstream from the mountain front. The drainage area, measured as the number of upstream contributing cells, and the elevation are then extracted from the flow accumulation grid and DEM respectively. The long profile data are assembled for channel slope-drainage area analysis. Following extensive sensitivity analyses described below, the long profiles were first smoothed using a lowess routine and the smoothed channel gradients were plotted against drainage area in log-log space. These graphs shows a clear initial plateau where increases in drainage area do not correspond equally to decreases in channel gradients (fig. 3). Our data illustrates the lack of a covariance between slope and area for small drainage areas, but in our case, some of this is an artifact of the profile smoothing. Such headwater portions of a channel usually correspond to areas dominated by debris-flows and/or landslides where channel slope does not vary (Montgomery & Dietrich, 1988; Montgomery & *alii*, 1996; Sklar & Dietrich, 1998; Stock & Dietrich, 2003). For the Secchia River analysis we chose to run the analysis both including and not including the first data, in order to actually evaluate their influence on the resulting θ and k_s. For the remaining two datasets, the 14 main rivers and the 15 upstream channels, we chose to omit these headwater data and a linear regression analysis was applied to the remaining slope-area data in order to extract the k_s (Y intercept) and θ (slope of the line) parameters of the equations (3) and (4). The criteria used to

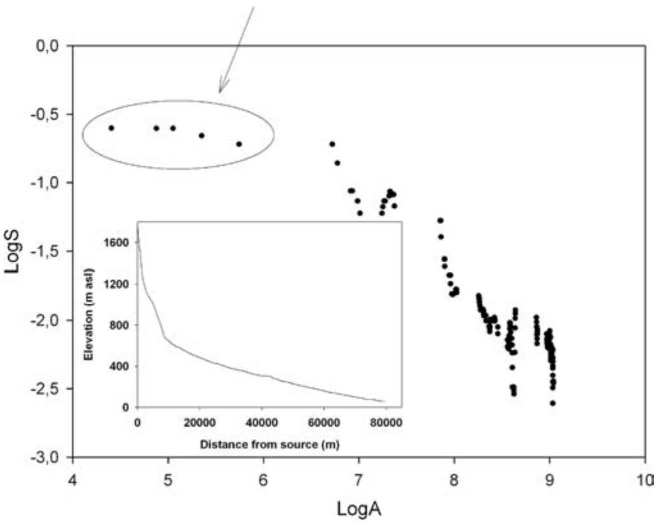

FIG. 3 - The longitudinal profile and the Log Slope vs. the Log Drainage Area profile of the Secchia River (logarithm was calculated from slope in degrees and drainage area in squared meters); the arrow shows the first headvalley data that were not taken into account for the regression analysis.

exclude the headwater data was based on the relative slope values for each channel. We excluded data where the channels were anomalously flat, using a threshold to remove cells that were not at least 1.4% steeper than the adjacent, downstream cell. In previous work, these «unfitting» data were found below the 0.1 km^2 drainage area threshold (Snyder & *alii*, 2000) or the 0.2 slope threshold (Sklar & Dietrich, 1998). In the study area rivers seems to have a highly variable drainage area threshold but, by averaging all values, we can define a lowest drainage area threshold of 2.54 km^2 and a lowest slope threshold of 0.19 (19%) which should draw the limit between debris flow-landslide dominated rivers and normal bedrock channels.

Rather than subjectively choosing specific reaches through which to regress, we consistently fit the best regression line through all of our data less the removed headwaters reaches described above. Regression lines that cross clear steps in the slope-area relationship suggest systematic downstream changes in concavity, caused by factors such as a migrating transient response, or a change in substrate or stream morphology. Regressions that cross such steps result in non-homoscedastic residuals and cannot be interpreted in the context of erosional theory (equation 1) and are really only a crude means of evaluating a general profile concavity (and steepness). The whole profile-regression approach is useful and warranted in our case, but we caution that such regressions cannot be interpreted in a theoretical context.

For each slope vs. drainage area plot a significance test was applied. The correlations were tested at the 1% level of significance with a Student t-test. The null hypothesis (H_0) was the total absence of correlation (Pearson coefficient equal to zero) between slope and drainage area and the results were considered significant only if the probability, p, to be wrong by rejecting H_0 was < 1%. If p exceeded 1% ($p>0.01$), the correlation was considered «non-significant».

RESULTS

Methodological development

We use the Secchia River longitudinal profile for testing the sensitivity of k_s and θ to various methods of long profile extraction and smoothing (fig. 3). In general some type of profile smoothing is a necessary step in avoiding inaccuracies that are common to all DEMs such as integer stepping that results in flat reaches and vertical knickpoints (see Snyder & *alii*, 2000 for a comparison of constant vertical interval and logarithmic binned averages approaches). Different smoothing techniques we investigated include running average, running median, bisquare, and lowess methods all available in the commercial software package SigmaPlot. The locally weighted, lowess routine works very well in eliminating integer stepping while still leaving the overall profile basically unchanged from the raw profile extracted from the DEM. In the lowess routine, each data point is reassigned a new value by application of a locally-weighted least-squares approach using a tricube weight function that takes into account a specified interval of the neighborhood data. In this analysis an interval that corresponds to the 5% of the number of values to be smoothed was found to be the best compromise between the attempt of being as close as possible to the original profile and the need of deleting integer stepping.

Different sampling intervals of 0.1, 0.2 and 0.4 km, in effect a low-pass filter of the smoothed profile, were extracted from the Secchia river profile. The test was applied to cases using the entire profile as well as the profile lacking the channel head data where slope and area do not correlate. In all cases (tab. 1) both θ and k_s show very little variation, within the same order of magnitude, in dependence of the sampling interval. There is a difference in the calculated q, and k_s between those profiles that include or exclude the channel head data. Both concavity and steepness are larger for the profiles that exclude the headwaters data than for those that include it. Considering that the two indexes do not show a correlation with the tested sampling intervals, we arbitrarily choose for the analysis of the main rivers a 0.4 km sampling interval. For the small channels, in order to have enough data to make the analysis statistically relevant, we worked with a sampling interval of 0.2 km.

Different drainage area downhill limits of 1000, 500 and 250 km^2 were tested both on the entire profile and the profile lacking the channel head data (tab. 1). In this case, the variation is more consistent, although still within the same order of magnitude and without a clear trend between the drainage area limits adopted and concavity or steepness. Also here, concavity and steepness tend to be larger for the profiles that exclude the headwaters data than for those that include it. Although there are not enough field data yet to exactly define the transitional zone (from bedrock to alluvial) for all rivers, we noticed that most Romagna rivers get to that point further down stream than the Emilia ones. At the light of the variability of concavity and steepness in relation to different drainage area downhill limits, a subjective but fixed 500 km^2 drainage area limit was adopted in the analysis of the trunk channel profiles. For our experience in the field, this limit could reasonably be upstream of the bedrock-alluvial transition zone in all the Romagna rivers but probably downhill, at least for some of the rivers, in the Emilia ones. For the small channels the downhill threshold, 60 km^2, was chosen to avoid a change in rock type. For both the main rivers

TABLE 1 - θ and k_s analysis of the Secchia River slope vs. the drainage area profile sampled at different intervals (A) and with different drainage area downhill limits (B), with (normal) and without (*italic* in grey background) the initial headvalley data where the river is dominated by debris-flow and landslide events

A - sampling distance	0.1 km	0.2 km	0.4 km	*0.1 km*	*0.2 km*	*0.4 km*
θ	0,47	0,48	0,49	*0,61*	*0,61*	*0,60*
ks	2,06	2,10	2,20	*3,18*	*3,21*	*3,20*
r^2	0,83	0,83	0,84	*0,85*	*0,86*	*0,84*

B - drainage area limit	1000 km^2	500 km^2	250 km^2	*1000 km^2*	*500 km^2*	*250 km^2*
θ	0,48	0,50	0,44	*0,61*	*0,76*	*0,76*
ks	2,06	2,26	1,86	*3,18*	*4,35*	*4,35*
r^2	0,83	0,81	0,78	*0,86*	*0,93*	*0,93*

and the small channels the first headwater data were not taken into account.

Analysis of trunk channel profiles

Fourteen selected large trunk river profiles were analyzed within drainage areas variously varying (especially in the upstream threshold) between 0.02 and 500 km^2 (tab. 2). Based on our sensitivity analyses carried out on the Secchia River profile, we sampled every 0.4 km along the river channels. The total length of the studied channels varies from a minimum of 29 km (the Taro River) to a maximum of 83 km (the Savena River). All profiles except the Savio showed high correlation coefficients so we determined a concavity (θ) and modeled k_s for the remaining 13 profiles. Concavity varies from a minimum value of 0.48 in the Panaro River to a maximum value of 1.42 in the Savena River (fig. 4 and fig. 5). The average θ value for all Emilia and Romagna rivers is 0,77. Besides the «anomalous» value of the Savena River, θ seems to have a geographical trend with higher values (more concave) in the Romagna province (from Bidente to Santerno, mean value of 0.83) and lower values in the Emilia province (from Savena to Ceno, mean value of 0.73).

The mean concavity of 0.77 is used to model profile steepness (k_s). Modeled profile steepness varies from a minimum of 7586 (Savena River) to a maximum of 36308 (Panaro River) with a mean value of 19884. Profile steepness is generally higher in the Emilia (mean = 24343) rather than Romagna (mean = 12645) Apennines.

Analysis of small channels

Fifteen additional smaller channels were analyzed for drainage areas between 0.02 and 60 km^2 in an effort to more carefully probe the influence of a single rock type on channel profile development (tab. 2). Within such a restricted drainage area interval, the analyzed channels have clearly incised into bedrock. Smoothed profiles were sampled every 0.2 km rather than 0.4 km so that more data points would be collected to model the shorter channels that ranged in length from 8 to 12 km. The results were tested with a Student t-test and only three channels were rejected at the 99% confidence interval: channels 4, 10, and 11.

The analysis of the thirteen remaining small channels yields values of concavity ranging from 0.25 (Rovigo Torrent) to 1.09 (Gotra Torrent) with a mean in the Emilia Apennines of 0.63 and a mean in the Romagna Apennines of 0.53 (fig. 4 and fig. 6). The average value for all of the small channels is 0.58 and using this mean value to model profile steepness reveals a mean k_s of 791 and 435 for the Emila and Romagna Apennines respectively.

DISCUSSION AND CONCLUSIONS

The analysis on a specific river, the Secchia, suggests that drainage area downhill limit and including or not the first headwater data, both influence the resulting values of concavity and steepness. This highlights how important it

TABLE 2 - θ and k_s analysis of the 14 main rivers and 15 upstream channels slope vs. the drainage area profiles. In *italics* and grey background are values derived from a fixed average θ; with an initial • are those values that were not taken into account because of their low r^2; mean values discussed in the paper are in **bold**

Name	ID #	Region	Length km	Area km²	rock type	lower 1/3 profile	samples distance	min A	min S	min ΔS	N data	θ	ks	r²	*θ*	*ks*	*r²*	signific. r²
Savio	1	Romagna	42	552	mixed (prevalent Marnoso-Arenacea)	bedrock	0,4	1,48	0,12	0,02	99	•-0,39	•17	0,52	*•0,77*	*•17378*	*0,04*	0,23
Bidente	2	Romagna	68	506	mixed (prevalent Marnoso-Arenacea)	bedrock	0,4	1,72	0,2	0,001	166	0,69	3548	0,76	*0,77*	*15849*	*0,75*	0,18
Rabbi	3	Romagna	75	541	mixed (prevalent Marnoso-Arenacea)	bedrock	0,4	1,68	0,16	0,03	184	0,9	117490	0,72	*0,77*	*10965*	*0,71*	0,17
Montone	4	Romagna	72	540	mixed (prevalent Marnoso-Arenacea)	bedrock	0,4	2,16	0,07	0,02	178	0,95	323594	0,73	*0,77*	*11482*	*0,70*	0,17
Lamone	5	Romagna	70	526	mixed (prevalent Marnoso-Arenacea)	bedrock	0,4	1,81	0,1	0,01	170	0,86	58884	0,83	*0,77*	*11749*	*0,82*	0,18
Santerno	6	Romagna	81	478	mixed (prevalent Marnoso-Arenacea)	bedrock	0,4	3,22	0,12	0,01	197	0,77	14791	0,66	*0,77*	*13183*	*0,66*	0,16
mean		**Romagna**	**73**	**524**	mixed (prevalent Marnoso-Arenacea)	bedrock						**0,83**				***12645***		
Standard Deviation												0,10				*1971*		
Savena	7	Emilia	83	254	mixed (prevalent Ligurian Nappe)	alluvial	0,4	5,40	0,06	0,01	202	1,42	1288249552	0,73	*0,77*	*7586*	*0,58*	0,16
Reno	8	Emilia	49	531	mixed (prevalent Ligurian Nappe)	alluvial	0,4	0,74	0,22	1E-04	124	0,59	631	0,93	*0,77*	*18621*	*0,84*	0,21
Panaro	9	Emilia	45	518	mixed (prevalent Ligurian Nappe)	alluvial	0,4	0,85	0,14	1E-04	113	0,48	148	0,83	*0,77*	*36308*	*0,52*	0,22
Secchia	10	Emilia	43	434	mixed (prevalent Ligurian Nappe)	alluvial	0,4	5,27	0,19	0,02	108	0,75	19952	0,9	*0,77*	*28184*	*0,90*	0,22
Parma	11	Emilia	57	493	mixed (prevalent Ligurian Nappe)	alluvial	0,4	2,83	0,18	0,01	137	0,63	1995	0,97	*0,77*	*28184*	*0,92*	0,20
Baganza	12	Emilia	75	361	mixed (prevalent Ligurian Nappe)	alluvial	0,4	1,55	0,15	0,01	184	0,65	2291	0,83	*0,77*	*23442*	*0,80*	0,17
Taro	13	Emilia	29	528	mixed (prevalent Ligurian Nappe)	alluvial	0,4	11,99	0,17	0,01	67	0,82	75858	0,95	*0,77*	*29512*	*0,94*	0,28
Ceno	14	Emilia	70	542	mixed (prevalent Ligurian Nappe)	alluvial	0,4	4,71	0,15	0,03	171	0,49	117	0,8	*0,77*	*22909*	*0,54*	0,18
mean		**Emilia**	**56**	**458**	mixed (prevalent Ligurian Nappe)	alluvial						**0,73**				***24343***		
Standard Deviation												0,30				*8584*		
Upper Savio	15	Romagna	10	59	dominant Marnoso-Arenacea	bedrock	0,2	0,17	0,33	1E-04	53	0,5	105	0,65	*0,58*	*355*	*0,64*	0,32
Bidente Pietrapazza	16	Romagna	11	43	dominant Marnoso-Arenacea	bedrock	0,2	0,14	0,34	0,01	61	0,61	646	0,69	*0,58*	*407*	*0,69*	0,30
Bidente Ridracoli	17	Romagna	9	50	dominant Marnoso-Arenacea	bedrock	0,2	1,42	0,25	0,03	55	0,76	9120	0,45	*0,58*	*427*	*0,42*	0,31
Bidente Corniolo	18	Romagna	10	45	dominant Marnoso-Arenacea	bedrock	0,2	0,63	0,42	0,02	47	0,64	1318	0,83	*0,58*	*525*	*0,82*	0,34
Upper Rabbi	19	Romagna	12	46	dominant Marnoso-Arenacea	bedrock	0,2	0,32	0,33	0,01	57	0,55	251	0,92	*0,58*	*437*	*0,92*	0,31
Upper Montone	20	Romagna	11	56	dominant Marnoso-Arenacea	bedrock	0,2	0,26	0,16	0,005	52	•0,29	288403	0,34	*0,58*	*•398*	*0*	0,32
Campigno	21	Romagna	12	40	dominant Marnoso-Arenacea	bedrock	0,2	0,27	0,17	0,015	61	0,40	26	0,81	*0,58*	*457*	*0,64*	0,30
Rovigo	22	Romagna	12	45	dominant Marnoso-Arenacea	bedrock	0,2	0,26	0,09	0,003	56	0,25	3	0,49	*0,58*	*•575*	*0*	0,31
mean		**Romagna**	**11**	**48**	dominant Marnoso-Arenacea	bedrock						**0,53**				***435***		
Standard Deviation												0,17				*56*		
Bratica	23	Emilia	10	47	dominant Ligurian and Sub-Ligurian U.	bedrock	0,2	0,46	0,24	0,002	46	0,44	98	0,89	*0,58*	*871*	*0,8*	0,34
Parma	24	Emilia	10	48	dominant Ligurian and Sub-Ligurian U.	bedrock	0,2	2,42	0,2	0,02	42	0,60	1122	0,58	*0,58*	*813*	*0,58*	0,36
Upper Baganza	25	Emilia	12	48	dominant Ligurian and Sub-Ligurian U.	bedrock	0,2	0,08	0,24	0,008	55	0,32	11	0,81	*0,58*	*•692*	*0,24*	0,31
Tarodine	26	Emilia	12	46	dominant Ligurian and Sub-Ligurian U.	bedrock	0,2	1,01	0,13	0,01	66	0,46	63	0,7	*0,58*	*468*	*0,66*	0,29
Manubiola	27	Emilia	11	55	dominant Ligurian and Sub-Ligurian U.	bedrock	0,2	3,47	0,15	0,03	55	0,6	1148	0,94	*0,58*	*813*	*0,94*	0,31
Gotra	28	Emilia	9	51	dominant Ligurian and Sub-Ligurian U.	bedrock	0,2	11,88	0,19	0,04	45	1,09	8128305	0,92	*0,58*	*1413*	*0,72*	0,35
Upper Ceno	29	Emilia	12	56	dominant Ligurian and Sub-Ligurian U.	bedrock	0,2	4,82	0,12	0,03	55	0,89	79433	0,67	*0,58*	*372*	*0,59*	0,31
mean		**Emilia**	**11**	**50**	dominant Ligurian and Sub-Ligurian U.	bedrock						**0,63**				***791***		
Standard Deviation												0,27				*367*		

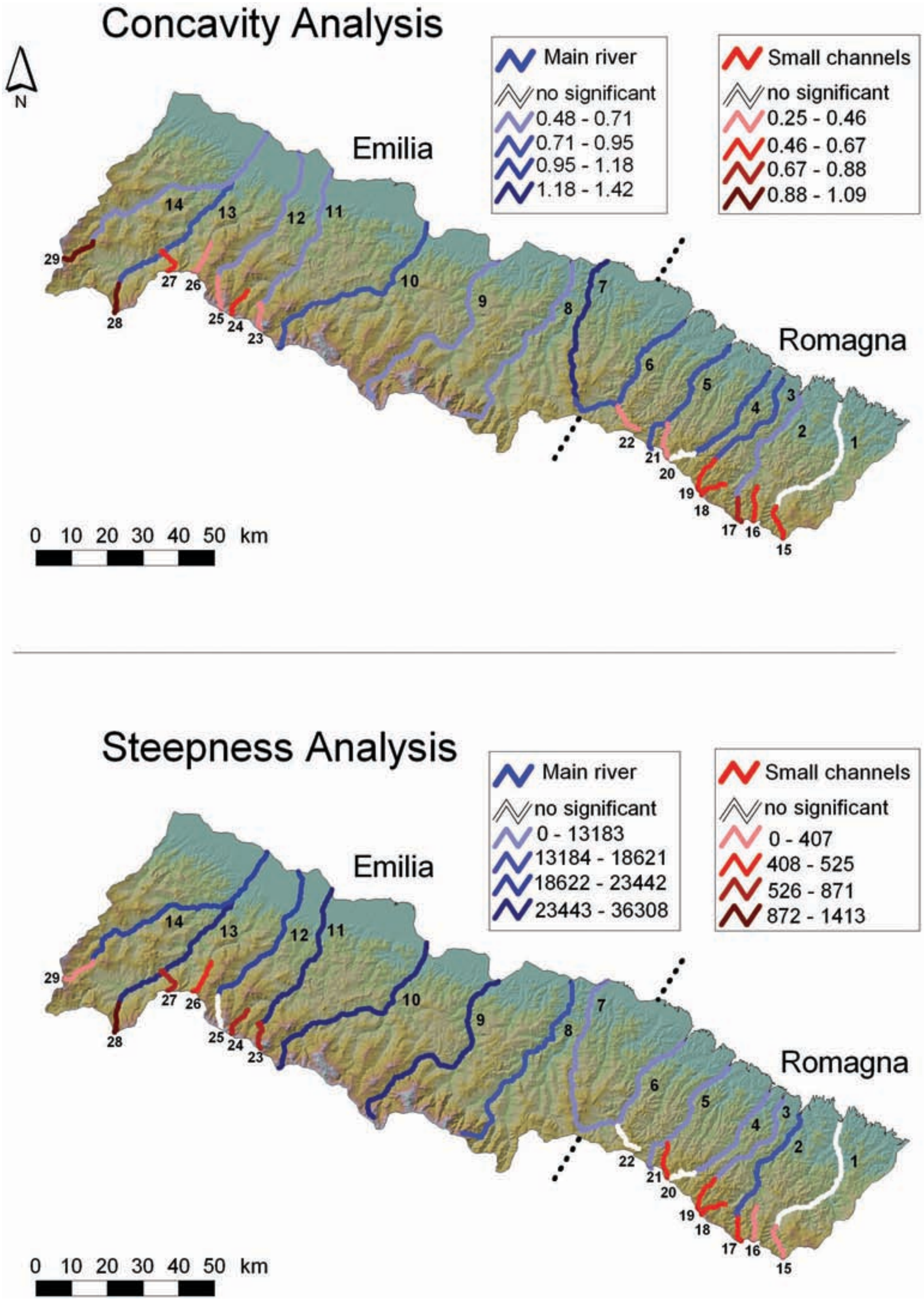

FIG. 4 - θ and modeled ks distribution in the 14 main rivers analyzed and the 15 upstream channels.

is to define a common methodology when focusing on a specific study area and how more attention should be paid on the different applied methods when comparing results from different works.

Raw and modeled profile concavity and steepness for the Northern Apennines cluster around a range of values that are in reasonable agreement with the range of theoretical values determined from stream power law modeling of detachment limited bedrock channels. Given the narrow range (same order of magnitude) of calculated profile concavity and steepness values, definitive interpretations in terms of rock type or tectonic controls are equivocal. Nevertheless, similar, recent studies report an equally narrow range of concavity and steepness values (Duvall et al., 2004) so if we take our mean concavity and steepness results for the Emilia and Romagna Apennines at face value, ignoring for the moment the poorly known standard errors, and incorporate these results

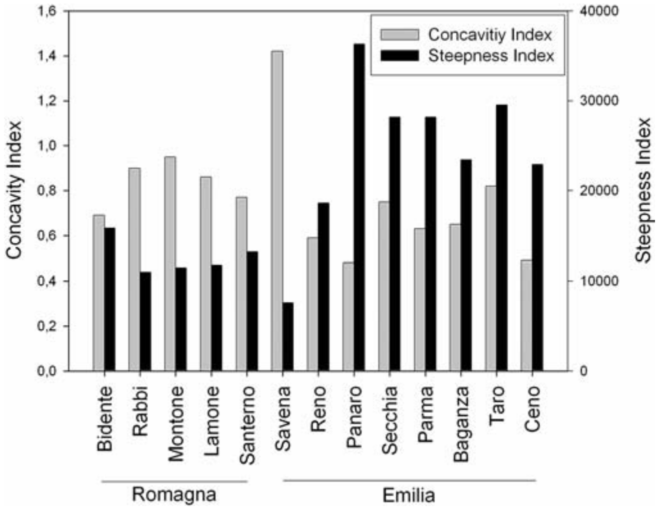

FIG. 5 - Plot of θ and modeled ks for the main rivers (values with no significant r^2 are not plotted).

with the local geologic setting, we can offer permissive interpretations.

Profile concavity and modeled steepness show some general trends that we can interpret in terms of rock type, tectonics, and the change in erosion process from detachment-limited (bedrock channel) to transport-limited (alluvial channel). Concavity tends to revolve around a relatively narrow range of values from 0.73 to 0.83 (means) for the channels of the large basins and 0.53 to 0.63 (means) for the channels of the small basins. The small channel results are interesting from the standpoint that they compare well to many other studies that have found concavities in the range of 0.4 to 0.6, values in good agreement with the theoretical predictions of equations (1) through (3) (Snyder & *alii*, 2000; Duvall & *alii*, 2004). The higher mean concavity values of the larger channel profiles may reflect the presence of transport limited processes in the larger channels because of alluvial reaches that are absent from the smaller, tributary bedrock channels. However, it is the Emilia Apennines that has common alluvial reaches in comparison to the Romagna Apennines, yet the concavity values of the Romagna Apennines trunk channels are higher than those for the Emilia Apennines. Modeled profile steepness does show consistent trends between the large and small channel profiles. In both cases, the profiles of the Emilia Apennines are steeper than the profiles of the Romagna Apennines.

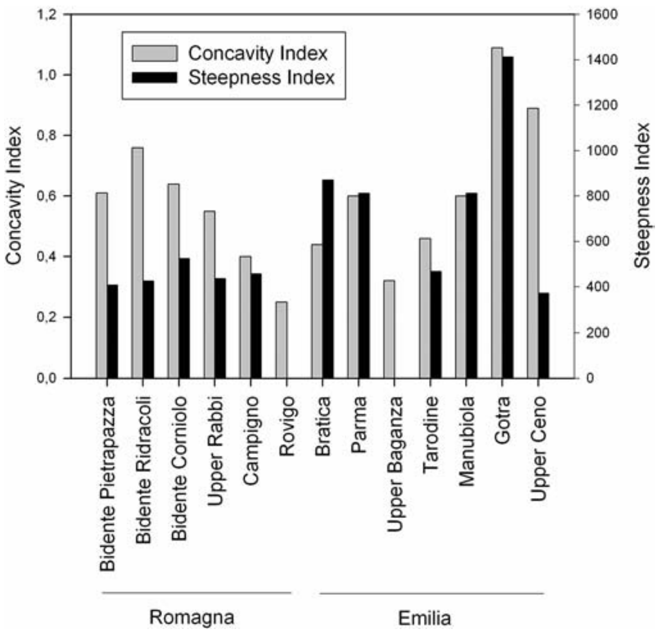

FIG. 6 - Plot of θ and modeled ks for the small channels (values with no significant r^2 are not plotted).

We are able to offer three possible interpretations, and criticisms of these results. The first possibility focuses solely on the profile steepness and looks at the variations of this index as a reflection of the differences in rock type from the Emilia to Romagna Apennines, while rock uplift is considered mostly invariant across the entire Northern Apennines. The general response to softer rock types is to have rivers that usually tend to reach their base level further upstream than rivers developed in harder rock type. This results in higher values of steepness. Profile steepness for both the large and small channels, higher in Emilia than in Romagna, is consistent with this response, assuming that the Emilia rocks are less resistant than the Romagna rocks. Nevertheless, differences in steepness values between the two regions are within the same order of magnitude, in contrast with what one could have predicted after Stock and Montgomery (1999). Thus, we suggest that our data are not nearly as supportive of a rock-type control only hypothesis as other recent studies (Snyder & *alii*, 2000, 2003; Duvall & *alii*, 2004). Completely lacking from our analysis is how channel width actually scales with drainage area. We have observed clear changes in channel width as a function of rock type, and particularly in the Emilia Apennines, we have observed channel width change as a function of the location of point alluvial sources, such as an active landslide. As long as channel width is allowed to freely vary as a function of rock type, then any change of profile gradient, expressed as concavity or steepness will be affected. The mixed signal that we get from concavity and steepness between the large trunk channels and the smaller headwater channels compels us to consider other possibilities rather than a simple rock-type-depending interpretation.

The second possibility takes into account both the concavity and steepness index variations and suggest that they reflect complicated variations in rock type, transport and detachment limited channel processes, and tectonics between the Emilia and Romagna Apennines. Channel concavity is increased by tectonics when the profile is being uplifted more strongly in its headwaters, than at its mouth (Kirby & Whipple, 2001). Studies of terrace long profiles along major rivers including the Taro (Benedetti & *alii*, 2003), the Reno (Amorosi & *alii*, 1996; Pazzaglia, unpublished data), and several Romagna streams (Boccaletti & *alii*, 2004) all conclude that the profiles diverge upstream, consistent with a model of more rock uplift in the headwaters in comparison to the mountain front. Incision rates increase from the mountain front towards the headwaters by a factor of 3-4 to an order of magnitude. Given this overall headward increase in incision rate, we would expect the profiles to be steeper and more concave where the tectonic forcing is stronger. Unfortunately we have not enough field data that would clearly show a higher incision rate, thus recent uplift, in the Romagna region rather than in the Emilia one. If we were to assume that the effect of tectonics on channel concavity and steepness is greater than the effects of rock type, we can interpret the small channel response as consistent with more recent rock uplift in the Emilia Apennines (higher concavity and steepness), and less recent rock uplift in the Romagna Apennines. These differences in rock uplift would have to overwhelm the expected counteracting effects of softer rocks and more alluvial channels in the Emilia Apennines and are not consistent with the fact that the Ligurian lid remains intact over this portion of the Apennines. In contrast, if we consider the trunk channel data only, the concavity values would argue for more recent uplift in the Romagna Apennines, consistent with the proposed recent stripping of the Ligurian rocks (Cerrina Feroni & *alii*, 2001), as long as we can attribute the steeper k_s values in the Emilia Apennines to the more alluvial nature of the channels there.

The third possibility focuses on both the concavity and steepness index of the small headwater channels and consider that tectonics are the only difference driving long profile forms. This interpretation argues for more rock uplift in the headwaters reach for the Emilia Apennines as compared to the Romagna Apennines, but probably similar rates of uplift for both regions in the direction of the mountain front. Such differential tilting to the headwaters to the southeast is consistent with higher mean elevations in the northwest, including the Mt. Cimone area. Unfortunately, such high mean elevations could also be explained by the exposure of the Macigno Formation in the headwaters of the Emilia Apennines and ignores the fact that there is solid evidence for active faults and active seismicity in the headwaters of the Romagna Apennines (Boccaletti & *alii*, 2004).

In summary, the slight differences in our values between the two portion of this sector of the Apennines, and the partially contrasting results we obtained from the large vs. the small channels analysis, all suggest that either the rock type and/or tectonic differences are not so relevant or there are other factors (i.e. channel width) that we did not take into account and that might have had a higher influence on the results. Nevertheless, if the subtle differences in our measured concavity and steepness values for small headwater and large trunk channels do speak to differences in tectonics or rock type for the Emilia and Romagna Apennines, we favor a mixed interpretation as opposed to an all rock-type or all tectonics control on long profile form.

Our results are encouraging from the perspective that they have served to solidify a consistent methodology and open the door to further investigations of a larger subset of small channels restricted to specific rock types to better ferret out the relative importance of rock type and tectonics on Northern Apennines landscape evolution.

REFERENCES

AMOROSI A., FARINA M., SEVERI P., PRETI D., CAPORALE L. & DI DIO G. (1996) - *Genetically related alluvial deposits across active fault zones: an example of alluvial fan; terrace correlation from the upper Quaternary of the southern Po Basin, Italy*. Sedimentary Geology, 102, 3-4, 275-295.

BARTOLINI C. & CARTON A. (1992) - *Cenni di geomorfologia. In Guide Geologiche Regionali: Appennino Tosco-Emiliano*. Società Geologica Italiana, BE-MA Editrice, 86-89.

BOCCALETTI M., BONINI M., CORTI G., GASPERINI P., MARTELLI L., PICCARDI L., SEVERI P. & VANNUCCI G. (2004) - *Carta sismotettonica della Regione Emilia-Romagna e Note Illustrative*. Servizio Geologico, Sismico, e dei Suoli, Regione Emilia Romagna, 60 pp, map scale 1:250,000.

BULL W.B. (1979) - *Threshold of critical power in streams*. Geological Society of America Bulletin, v. 90, 453-464.

BENEDETTI L., TAPPONNIER P., GAUDEMER Y., MANIGHETTI I. & VAN DER WOERD J. (2003) - *Geomorphic evidence for an emergent active thrust along the edge of the Po Plain: the Broni-Stradella fault*. Journal of Geophysical Research, v. 108, doi:10.1029/2001JB001546.

CERRINA FERONI A., LEONI L., MARTELLI L., MARTINELLI P., OTTRIA G. & SARTI G. (2001) - *The Romagna Apennines, Italy: an eroded duplex*. Geological Journal, v. 36, 39-54.

DUVALL A., KIRBY E. & BURBANK D. (2004) - *Tectonic and lithologic controls on bedrock channel profiles and processes in coastal California*. Journal of Geophysical Research, 109, F03002, doi 10.1029/2003 JF000086.

ELLIS M., CAI X. & ANDERSON R.S. (1997) - *The making of a graded river profile as a coupled bedload and sediment transport process*. EOS (AGU abs).

FEDERICI P.R. & TELLINI C. (1983) - *La geomorfologia dell'Alta Val Parma (Appennino Settentrionale)*. Rivista Geografica Italiana, 90, 393-428.

FLINT J.J. (1974) - *Stream gradient as a function of order, magnitude and discharge*. Water Resources Research, v. 10, 969-973.

HACK J.T. (1957) - *Studies of longitudinal stream profiles in Virginia and Maryland*. U.S. Geological Survey Professional Paper, 294-B, 45-97.

HOWARD A.D. (1971) - *Problems in interpretation of simulation models of geologic processes*. In: Quantitative Geomorphology, Some Aspects and Applications, edited by M. Morisawa, 61-82.

HOWARD A.D. & KERBY G. (1983) - *Channel changes in badlands*. Geological Society of American Bulletin, v. 94, 739-752.

HOWARD A.D., SEIDL M.A. & DIETRICH W.E. (1994) - *Modeling fluvial erosion on regional to continental scale*. Journal of Geophysical Research, v. 99, 13,971-13,986.

HURTREZ J.-E., LUCAZEAU F., LAVÉ J. & AVOUAC J.-P. (1999) - *An investigation of the relationships between basin morphology, tectonic uplift, and denudation from the study of an active fold belt in the Siwalik Hills, central Nepal*. Journal of Geophysical Research, v. 104, 12796-12799.

KIRBY E. & WHIPPLE K. (2001) - *Quantifying differential rock-uplift rates via stream profile analysis*. Geology, 29, 415-418.

KNEUPFER P.L.K. (1994) - *River long profiles in active orogens, Taiwan and new Zealand; how much tectonic signal?* Eos, Transactions, American Geophysical Union, 75, n. 44, 288 pp.

LEOPOLD L.B., WOLMAN G. & MILLER J.B. (1964) - *Fluvial Processes In Geomorphology*, W. H. Freeman and Company, San Francisco, California, 522 pp.

MACKIN J.H. (1948) - *Concept of the graded river*. Geological Society of America Bulletin, 59, 463-512.

MAYER L. (2000) - *Application of digital elevation models to macroscale tectonic geomorphology*. In: Summerfield M.A. (ed.), «Geomorphology and Global Tectonics», John Wiley and Sons, London, 15-27.

MOGLEN G.E. & BRAS R.L. (1995) - *The effect of spatial heterogeneities on geomorphic expressions in a model of basin evolution*. Water Resources Research, 31, 2613-2623.

MONTGOMERY D.R. & DIETRICH W.E. (1988) - *Where do channels begin?* Nature, 336, 232-234.

MONTGOMERY D.R., ABBE T.B., BUFFINGTON J.M., PETERSON N.P., SCHMIDT K.M. & STOCK J.D. (1996) - *Distribution of bedrock and alluvial channels in forested mountain drainage areas*. Nature, 381, 587-589.

OHOMORI H. (1991) - *Change in the mathematical function type describing the longitudinal profile of a river through an evolutionary process*. Journal of Geology, 99, 97-110.

PAZZAGLIA F.J., GARDNER T.W. & MERRITTS D. (1998) - *Bedrock fluvial incision and longitudinal profile development over geologic time scales determined by fluvial terraces*. In: «Rivers over rock: Fluvial processes in bedrock channels», American Geophysical Union, Geophysical Monograph Series, 107, 207-235.

REGIONE EMILIA-ROMAGNA, SERVIZIO METEOROLOGICO REGIONALE (1995) - *I numeri del Clima: temperature, precipitazioni e vento. Tavole climatologiche dell'Emilia-Romagna 1951-1994*. Promodis Italia, Brescia, 300 pp.

ROE G., MONTGOMERY D.R. & HALLET B. (2002) - *The effects of orographic precipitation variations on steady-state river profiles*. Geology, 30, 143-146.

SCHUMM S.A. & LICTHY R.W. (1965) - *Time, space, and causality in geomorphology*. American Journal of Science, 263, 110-119.

SEIDL M.A. & DIETRICH W.E. (1992) - *The problem of channel erosion into bedrock*. Catena, Supplement, 23, 101-124.

SINHA S.K. & PARKER G. (1996) - *Causes of concavity in longitudinal profiles of rivers*, Water Resources Research, 32, 1417-1428.

SKLAR L. & DIETRICH W.E. (1998) - *Longitudinal river profiles and bedrock incision models: Stream power and the influence of sediment supply*. In: Tinkler K.J. & Wohl E.E. (eds.), «Rivers over rock: Fluvial processes in bedrock channels», American Geophysical Union Geophysical Monograph, 107, 237-260.

SLINGERLAND R., WILLETT S.D. & HOVIUS N. (1998) - *Slope-area scaling as a test of fluvial bedrock erosion laws*. EOS, Transaction of the American Geophysical Union 79, F358 (Fall Meet. Suppl.).

SNYDER N.P., WHIPPLE K.X., TUCKER G.E. & MERRITTS D.J. (2000) - *Landscape response to tectonic forcing: digital elevation model analysis of stream profiles in the Mendocino triple junction region, northern California*. Geological Society of American Bulletin, 112, 1250-1263.

SNYDER N.P., WHIPPLE K.X., TUCKER G.E. & MERRITTS D.J. (2003) - *Importance of a stochastic distribution of floods and erosion thresholds in the bedrock river incision problem*. Journal of Geophysical Research, B, Solid Earth and Planets, 108, 10.1029/2001JB001655.

STOCK J.D. & MONTGOMERY D.R. (1999) - *Geologic constraints on bedrock river incision using the stream power law*. Journal of Geophysical Research, 104, B3, 4983-4993.

STOCK J.D. & DIETRICH W.E. (2003) - *Valley incision by debris flows: Evidence of topographic signature*. Water Resourses Research, 39, 1089, doi:10.1029/2001WR001057.

TUCKER G.E. & WHIPPLE K.X. (2002) - *Topographic outcomes predicted by stream erosion models: Sensitivity analysis and intermodel comparison.* Journal of Geophysical Research, 107 (B9), 2179, doi: 10.1029/2001JB000162.

WHIPPLE K.X. & TUCKER G.E. (1999) - *Dynamics of the stream-power river incision model: implications for height limits of mountain ranges, landscape response timescales, and research needs*. Journal of Geophysical Research, 104, 17,661-17,674.

WILLGOOSE G. (1994) - *A statistic for testing the elevation characteristics of landscape simulation models*. Journal of Geophysical Research, 99 B7, 13,987-13,996.

ZAPROWKI B.J., PAZZAGLIA F.J. & EVENSON E.B., in press (2005) - *Climatic influences on long profile concavity and river incision*. Journal of Geophysical Research - Earth Surface, 110, XXXXXX, doi: 10.1029/2004FJP0000138.